Modeling Barton Springs Segment of the Edwards Aquifer using MODFLOW-DCM

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Abstract

The Barton Springs segment of the Edwards Aquifer is the sole source of water supply for about 45,000 people in and immediately south of Austin, Texas. For water management purposes, it is important to be able to predict the availability of groundwater in response to future development and potential droughts. The Edwards Aquifer comprises heterogeneous carbonate rock strata that have developed a well-connected network of karst conduits. This karstic geological structure makes aquifer characterization and groundwater modeling challenging. Previously, Scanlon et al. (2001) developed a two-dimensional groundwater model for the Barton Springs segment using MODFLOW. The karst conduits were not explicitly represented in the model. Instead, the study area was divided into 9 zones for which the transmissivities were obtained through calibration. We revisit Scanlon et al.’s work in this study by using MODFLOW-DCM, a MODFLOW module developed to represent flow through karstic aquifers. MODFLOW-DCM adopts a dual-conductivity approach in which the aquifer is conceptualized as being composed of two interacting flow systems, i.e., the background matrix and the karst conduits. This approach allows karst aquifers to be modeled as coupled systems, and thus allows aquifer dynamics related to karst conduit flows to be accurately simulated. Our preliminary results show improved matching of both water level measurements and spring discharges records.

Introduction

The Edwards Aquifer is one of the largest and most important karst aquifer systems in the United States. The karstic Edwards Aquifer not only serves as a main source of water supply for about 1.7 million users in south central Texas, US, but also habitat for over forty aquatic species, some of which are listed as endangered species (Longley, 1986; Edwards et al., 1989). As demand for water continues to rise, there are increasing concerns about the reducing groundwater availability and its impact on the welfare of endangered species and the regional economy. It is thus imperative to advance the current characterization and understanding of the Edward Aquifer by providing better tools for groundwater resource management.

Numerous studies have been carried out in the past on the Edwards Aquifer. For example, Pearson and Rettman (1976) estimated the age of water at several springs of the Edward Aquifer using tritium isotopes. Slade et al. (1985) developed a two-dimensional groundwater flow model for the Barton Springs segment of the Edwards Aquifer. Kuniansky (1994) set up a finite-element model of the Edwards and Trinity Aquifers. Kuniansky’s finite-element model was designed to incorporate the geologic and hydrologic conditions that affect groundwater flow. Barrett and Charbeneau (1996) developed a lumped-parameter model to predict the impacts of urban development on the quantity of water in the Barton Springs segment of the Edwards Aquifer. Kuniansky et al. (2001) estimated travel times along selected flow paths of the Edwards Aquifer using simulated groundwater levels from a quasi three-dimensional finite-element model. In the quasi three-dimensional model, groundwater flow is simulated as horizontal within two model layers, with vertical leakage occurring between the layers. Kuniansky et al. (2001) concluded that their estimation of flow direction and Darcy flux along selected paths is reasonable, but their estimates of pore velocity and travel time were less conclusive due to the limited information on the effective aquifer thickness and the effective porosity. Scanlon et al. (2001) developed a two-dimensional finite-difference model for the Barton Springs segment of the Edwards Aquifer. The model was intended to evaluate groundwater availability, and predict water levels and spring flow in response to increased estimated by calibrating the steady-state model using trial and error, and automated inverse methods. Good agreement was found between measured and simulated spring discharge at Barton springs and water levels in many of the monitoring wells.

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In all the above-mentioned studies, the aquifer was essentially modeled with a single continuum approach in which no distinction was made between karst conduits and rock matrix. Instead, effective parameters were assigned to each numerical cell. It is well known that flow in karst aquifers occurs at multiple spatial and temporal scales. In the case of the Edwards Aquifer, dissolution of limestone forms caverns that are much more conductive than the rock matrix (Kuniansky et al., 2001). Failure to recognize this multi-scale nature of karst aquifers can lead to a model with incorrect structure. Even if such a model can be calibrated with various measurement data, it may still produce inaccurate results when used for prediction purposes. In general, a model based on the single continuum approach may not simultaneously fit the water level data and spring discharge records to a satisfactory degree (Painter et al., 2004) due to the large contrast between hydraulic properties (conductivity and storage coefficient) in the rock matrix and in conduits. While flow regime in the former is laminar, flow in latter often falls in the turbulent regime.

Painter et al. (2004) recently applied a dual-conductivity approach to modeling flow in karst aquifers. The dual-conductivity approach enables one to model a karst aquifer as a coupled system of conduits and rock matrix, with the coupling term being the cross flow between conduits and matrix. The advantages of this dual conductivity approach are: (a) it makes explicit representation of conduit geometry possible; (b) it provides an additional degree of freedom (via the cross-flow or exchange term) for simultaneous fitting of spring discharge and well water levels; and (c) it allows one to apply different flow laws for conduit and matrix flows. Based on the dual conductivity concept, Painter et al. (2004) developed an add-in module, the Dual Conductivity Module (DCM), for the groundwater flow simulation package MODFLOW (Harbaugh and McDonald, 1996; Harbaugh et al., 2000).

In this work, we use MODFLOW-DCM to model the Barton Springs segment of the Edwards Aquifer. The Barton Springs segment is chosen because its geological structure, especially karst conduits, has been extensively characterized via tracer studies (Hauwert et al., 1998, 2001). The single continuum, MODFLOW model developed by Scanlon et al. (2001) provided an excellent starting point for our application. We demonstrate the effectiveness of MODFLOW-DCM for modeling karst aquifers by using the same model configuration as that was given in Scanlon et al. (2001), but replacing the single continuum representation of Scanlon et al. with a dual-conductivity representation.

Background

Description of the Study Area

Unless otherwise noted, most of the description provided herein is based on the report of Scanlon et al. (2001). The Barton Springs segment of the Edwards Aquifer includes parts of Travis and Hays counties. Surface elevations in the study area range from about 320 m above mean sea level (amsl) in the southwest to about 76 m amsl along the east margin (Figure 1). The study area is in the subtropical humid climate zone. Annual precipitation ranges from 279 to 1,651 mm (1860 through 2000), based on records from a NOAA station located north of the study area. Long-term mean annual precipitation is 851 mm.

About 80 percent of the Barton Springs segment is unconfined and the remainder is confined. The recharge mainly occurs in the outcrop area in the north. The western model boundary is Mount Bonnell fault, which acts as a no-flow boundary; the southern boundary is a groundwater divide along Onion Creek; the eastern boundary is the saline-water line defined by the 1,000 TDS (Total Dissolved Solids) contour; and the northern boundary is the Colorado River.

The study area includes five major drainage basins. The primary source of recharge is seepage from streams crossing the outcrop area. The estimated area of the recharge zone is about 233 km$^2$. Most flow in the aquifer discharges in Barton Springs. The average spring discharge from 1917 to 1998 was 1.50 m$^3$/s. The discharge of the nearby Cold Springs is only a small percentage of the Barton Springs (as low as 4 percent).
Based on aquifer test data, effective values of hydraulic conductivity were found to range from 0.12 to 22.95 m/d. The range of specific yield was found to be 5E-3 to 0.06, and the range of specific storage was found to be from 3.28E-6 to 9.51E-2 m⁻¹. Tracer studies by Hauwert et al. (1998, 2001) indicated that groundwater flow can be as fast as 1.13E4 m/d along major pathways. Conduit flow paths inferred by Hauwert et al. (1998, 2001) using the tracer study results are illustrated in Figure 2.

**The Single Continuum Model**

Scanlon et al. (2001) developed a two-dimensional MODFLOW model for assessing different water management scenarios. The model grid (Figure 1) consists of one layer that has 120 rows and 120 columns. The model layer was assigned as the confined/unconfined type in MODFLOW. Model rows were aligned parallel to the strike of the Edwards. The cells outside of the model boundary, as well as the cells with thickness less than 6.10 m, were made inactive, resulting in a total of 7,043 active cells.

The drainage package of MODFLOW was used to represent Barton Springs and Cold Springs. The elevations of Barton Springs and Cold Springs are 132 m amsl and 131 m amsl, respectively. A high drain conductance of 9.29E4 m²/d was used to allow unrestricted discharge of water.

The spatial distribution of recharge among the streams and in the inter-stream area settings was based on the average recharge for a 20-year record. The total amount of recharge was reduced to match the average spring discharge in Barton Springs and Cold Springs.

The distribution of hydraulic conductivity was estimated by Scanlon et al. (2001) using a combination of trial and error and automated inverse approaches. The trial and error procedure indicated that there are 9 zones of hydraulic conductivity ranging from 0.30 to 304.80 m/d. An inverse code was then used to improve the manual estimates by attempting to reduce the root-mean error between simulated and observed water levels. With the final zonal distribution of hydraulic conductivity, the errors between measured and simulated heads were generally low except in the southwest area. The simulated discharge was 1.47 m³/s at Barton Springs and 7.93E-2 m³/s at Cold Springs.

After calibrating the steady-state model, Scanlon et al. (2001) ran a transient model for a 10-year period (1989 through 1998) by using monthly recharge and pumpage rates. Peak discharges at the Barton Springs simulated with the single continuum model were at times underestimated. The single continuum model predicted that the slope of the simulated springflow recession hydrograph was more gradual than that of the measured recession, and the timing of the simulated spring hydrograph was later than that of the measured data. The water levels in some wells could not be reproduced in the simulation because they are located in the vicinity of suspected conduits.

As mentioned in the Introduction, one of the advantages of the dual-conductivity representation lies in its ability to better reproduce the hydraulic dynamics of a karst system. We are thus interested in testing the effectiveness of MODFLOW-DCM through revisiting Scanlon et al.’s work. We report our preliminary findings in the following sections.

**Numerical Approach**

The theoretical background of the dual-conductivity approach adopted here is documented in detail in Painter et al. (2004) and summarized here. The DCM module is based on the Layer Property Flow (LPF) module of MODFLOW 2000 (Harbaugh et al. 2000). The current implementation of DCM assumes one model layer for conduit network and one model layer for rock matrix. It is important to note that in DCM the conduit network layer and the rock matrix layer are two physically interacting layers. The matrix layer is assumed to be pervasive. The conduit network layer can be specified as sparse or pervasive depending on the nature of the conduit system being modeled. The thickness of each layer can vary in space, but the conduit layer should always be contained in the matrix layer. Mass exchange between the conduits and rock matrix is governed by a linear exchange term. As a result, the current DCM can be considered as a 2.5-dimensional model.
Similar as in the LPF module, spatial distributions of the hydraulic conductivity and storativity (or special yield) can be specified via either zone or matrix methods. In practice, the hydraulic parameters and the linear exchange term need to be obtained through model calibration. The hydraulic parameters are expected to be effective properties of the conduit and its adjacent areas. The vertical leakage term is no longer needed. Because conduits are allowed to be partially filled or dry, upstream weighting is used in place of the original MODFLOW rewetting scheme for numerical stability.

The flow regime in conduits can be either addressed by the Darcy equation (for laminar flow) or Darcy Weisbach equation (for turbulent flow). In a future version, a scheme will be introduced to automate the transition from laminar flow regime to turbulent flow regime.

**Numerical Model**

When converting Scanlon et al.’s (2001) original two-dimensional model into a DCM model, we used the same elevation and thickness of the original single-layer, single continuum model for the matrix layer. We placed conduits underneath the six losing (i.e., recharge) streams and the two major groundwater pathways as inferred from Hauwert et al.’s (2001) tracer study, which is shown in Figure 2. Our rationale was that conduits or conduit-like zones directly contributed to the high conductivity in those areas. The resulting discrete conduit network is plotted in Figure 3, where the conduits are labeled in dark color and the matrix is in gray color. For simplicity, we did not include Cold Springs in this study because it only accounts for a small percentage of the total spring discharge. We assumed in this preliminary study that the conduit layer has the same thickness as the matrix layer. It is possible that in some areas the vertical extent of the conduit is smaller. Refinement will be possible when additional characterization data for the study area become available.

High values of hydraulic conductivity were assigned to the conduits. The two main conduits were assigned a value of 1,067 m/d, the conduits under the losing streams were assigned a value of 61 m/d, and the area immediately adjacent to Barton Springs was assigned a value of 610 m/d. The zonal pattern of hydraulic conductivity as determined by Scanlon et al. (2001) was retained in the DCM matrix layer, with values of hydraulic conductivity ranging from 0.30 to 9.14 m/d. The linear exchange term was set at 1.00E-3 d⁻¹ for the conduits under the losing streams and 5.00E-4 d⁻¹ for the main conduits and the Barton Springs discharge area. In this work, selection of most of the parameters was based on manual calibration using the steady-state discharge of Barton Springs as a criterion.

**Numerical Results**

The steady-state hydraulic head distributions in the conduit and matrix layers are shown in Figure 4 as shaded density plots. Hydraulic head increases from 131.06 m amsl at Barton Springs to more than 243.84 m amsl in the western edge of the recharge zone. Hydraulic gradients are generally low in the confined region due to the influence of the larger major conduits located in the confined region. Hydraulic gradients are larger in the recharge zone where conduits are smaller. These features are in general agreement with water elevations measured during the 1992 period. Contours of observed water elevations (Scanlon et al., 2001, 2003) from this period are shown for comparison in Figure 5.

A 1-year transient simulation was also performed. Hydraulic heads calculated with the steady-state model were used as the initial condition. The simulation used 1-month recharge periods. The recharge multipliers are shown in Figure 6. These are intended to approximate the recharge for the year 1989. The time steps ranged from 1 to 3 days, depending on the recharge period. Pumping was constant during the simulation and the same as in the steady-state model.

Shaded density plots of the calculated hydraulic heads in the conduit and matrix system at the end of month 5 are shown in Figure 7. Month 5 is a period of high recharge. The conduit head is higher than the matrix system head during this period, and the flow is from the conduit to the matrix system. Groundwater ridges develop in the matrix system in the recharge zone and to a lesser extent in the confined zone.
Similar shaded density plots at the end of the 1-year simulation are shown in Figure 8. The last 5 months of the simulation have no recharge and the hydraulic heads at the end of the simulation period are considerably lower than in Figure 7. Moreover, significant groundwater troughs have developed along the conduit positions.

The spring-flow hydrograph and a water-level hydrograph for two wells are shown in Figure 9. The spring flow responds very quickly to changes in recharge (Figure 9A). For the transient simulation, the springs respond within hours of a change in recharge. This response time can be adjusted by changing specific yield in the conduit system or the conduit/matrix exchange term. The water-level hydrograph for a location near Barton Springs is very unresponsive (subdued) and changes by less than 0.61 m during the 1-year simulation (Figure 9C). At the location near the southernmost conduit (Figure 9B), the water level responds slower than the springs and fluctuates by about 3 m during the simulation.

The general responses of the hydrographs are roughly consistent with observed hydrographs. Of particular interest is the relatively subdued response for water levels near the conduit head (Figure 9C). In the single-continuum MODFLOW simulation (Scanlon et al., 2001), the water level in that location is overly responsive to changes in recharge and changes by about 15 m during the 1989 period. The less responsive behavior in the DCM simulations is encouraging and suggests that the additional flexibility inherent in the DCM approach may allow better representation of dynamic behavior in karst aquifers. In addition, the DCM model was able to simulate groundwater troughs and ridges commonly observed near conduits in karst aquifers.

Conclusions

MODFLOW-DCM (Painter et al, 2004), an extension of MODFLOW for modeling conduit flows, was used in this work to model the Barton Springs segment of the Edwards Aquifer. The model structure and boundary conditions are based on the single continuum model of Scanlon et al. (2001) for the same study area. Property assignments for the DCM matrix layer and the conduit network were based on the properties of their single layer continuum model, the tracer study results of Hauwert et al. (1998, 2001), and hydrogeological interpretation of the study area.

Specific segments of the conduit network were aligned with recharge streams and two large conduit flow features inferred from tracer test results. The matrix layer was assigned hydraulic properties similar to those of the lower permeable zones assigned by Scanlon et al. (2001). Hydraulic properties appropriate for rapid conduit flow were assigned to the conduit network. The property values for the matrix/fracture interaction term were determined during the steady-state and transient calibration process.

Initial application of the DCM package to the Barton Springs segment of the Edwards Aquifer demonstrates that the approach can be successfully used in practice. One-year transient simulations of the Barton Springs segment produced flashy spring response, relatively subdued water level response, and transient groundwater troughs. These phenomena are all consistent with the observed behavior of the aquifer.

References


Acknowledgment

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Figure 1. The finite-difference grid used in Scanlon et al.’s MODFLOW model. The solid line delineates the model boundary (from Scanlon et al., 2001).
Figure 2. The figure shows dye injection locations (red circles) and inferred flow paths (solid pink lines and dotted pink lines). The locations of dye detections are shown as green circles and represent wells and springs (from Hauwert et al., 2001).
Figure 3. The discrete conduit network created for Barton Springs segment of Edwards Aquifer. Conduits are placed under losing streams and the major groundwater pathways. The $x$ and $y$ axes represent column and row numbers in the MODFLOW-DCM model. Each grid cell has dimensions 1,000 ft × 500 ft (304.8 m × 152.4 m).
Figure 4. Simulated hydraulic heads in the conduit (upper plot) and matrix (lower plot) layers for the Barton Springs steady-state model. Each grid cell has dimensions 1,000 ft $\times$ 500 ft (304.8 m $\times$ 152.4 m).
Figure 5. Contours of measured water elevations (in ft) in the Barton Springs Segment of the Edwards Aquifer during the 1992 period (Scanlon et al., 2001, 2003).
Figure 6. Monthly recharge multiplier used in the 1-year transient model of the Barton Springs Segment. Values shown are relative to the long-term average.
Figure 7. Simulated hydraulic heads in the conduit (upper plot) and matrix (lower plot) layers at the end of month 5 in the Barton Springs 1-year transient simulation.
Figure 8. Simulated hydraulic heads in the conduit (upper plot) and matrix (lower plot) layers at the end of the Barton Springs 1-year transient simulation.
Figure 9. Results from the Barton Springs transient simulation. Plot A is the spring-flow hydrograph at the Barton Springs; Plot B and C are two water-level hydrographs that correspond to a location near the southernmost conduit and a location near the Barton Springs.